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CHAPTER 25—120 to 0 Ma tectonic evolution of the southwest Pacific and analogous geological evolution of the 600 to 220 Ma Tasman Fold Belt System

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We review the tectonic evolution of the SW Pacific east of Australia from ca 120 Ma until the present. A key factor that developed early in this interval and played a major role in the subsequent geodynamic history of this region was the calving off from eastern Australia of several elongate microcontinental ribbons, including the Lord Howe Rise and Norfolk–New Caledonia Ridge. These microcontinental ribbons were isolated from Australia and from each other during a protracted extension episode from *ca* 120 to 52 Ma, with oceanic crust accretion occurring from 85 to 52 Ma and producing the Tasman Sea and the South Loyalty Basin. Generation of these microcontinental ribbons and intervening basins was assisted by emplacement of a major mantle plume at 100 Ma beneath the southern part of the Lord Howe Rise, which in turn contributed to rapid and efficient eastward trench rollback. A major change in Pacific plate motion at *ca* 55 Ma initiated east-directed subduction along the recently extinct spreading centre in the South Loyalty Basin, generating boninitic lithosphere along probably more than 1000 km of plate boundary in this region, and growth of the Loyalty–D'Entrecasteaux arc. Continued subduction of South Loyalty Basin crust led to the arrival at about 38 Ma of the 70–60 million years old western volcanic passive margin of the Norfolk Ridge at the trench, and west-directed emplacement of the New Caledonia ophiolite. Lowermost allochthons of this ophiolite are Maastrichtian and Paleocene rift tholeiites derived from the underthrusting passive margin. Higher allochthonous sheets include a poorly exposed boninitic lava slice, which itself was overridden by the massive ultramafic sheets that cover large parts of New Caledonia and are derived from the colliding forearc of the Loyalty–D'Entrecasteaux arc. Post-collisional extensional tectonism exhumed the underthrust passive margin, parts of which have blueschist and eclogite facies metamorphic assemblages. Following locking of this subduction zone at 38–34 Ma, subduction jumped eastward, to form a new west-dipping subduction zone above which formed the Vitiaz arc, that contained elements which today are located in the Tongan, Fijian, Vanuatu and Solomons arcs. Several episodes of arc splitting fragmented the Vitiaz arc and produced first the South Fiji Basin (31–25 Ma) and later (10 Ma to present) the North Fiji Basin. Collision of the Ontong Java Plateau, a large igneous province, with the Solomons section of the Vitiaz arc resulted in a reversal of subduction polarity, and growth of the Vanuatu arc on clockwise-rotating, older Vitiaz arc and South Fiji Basin crust. Continued rollback of the trench fronting the Tongan arc since 6 Ma has split this arc and produced the Lau Basin–Havre Trough.

This southwest Pacific style of crustal growth above a rolling-back slab is applied to the 600–220 Ma tectonic development of the Tasman Fold Belt System in southeastern Australia, and explains key aspects of the geological evolution of eastern Australia. In particular, collision between a plume-triggered 600 Ma volcanic passive margin and a 510–515 Ma boninitic forearc of an intra-oceanic arc had the same relative orientation and geological effects as that which produced New Caledonia. A new subduction system formed probably at least several hundred kilometres east of the collision zone and produced the Macquarie arc, in which the oldest lavas were erupted *ca* 480 Ma. Continued slab rollback induced regional extension and the growth of narrow linear troughs in the Macquarie arc, which persisted until terminal deformation of this fold belt in the late-Middle to Late Devonian. A similar pattern of tectonic development generated the New England Fold Belt between the Late Devonian and Late Triassic. Parts of the New England Fold Belt have been broken from Australia and moved oceanward to locations in New Zealand, and on the Lord Howe Rise and Norfolk–New Caledonia Rise, during the post-120 Ma breakup. Given that the Tasman Fold Belt System grew between 600 and 220 Ma by crustal accretion like the southwest Pacific since 120 Ma, facing the open Pacific Ocean, we question whether the eastern (Australia–Antarctica) part of the Neoproterozoic Rodinian supercontinent was joined to Laurentia.

KEY WORDS: Australia, Lachlan Fold Belt, southwest Pacific, subduction, Tasman Fold Belt, tectonics.

INTRODUCTION

The Tasman Fold Belt System makes up most of the crust of eastern Australia and consists of three broadly north–south-oriented fold belts (Figure 1) that record a continuing history of crustal growth from 600 Ma until at least the Triassic. Taking into account the 800–600 Ma Adelaide Rift Complex further west, this eastward younging of the

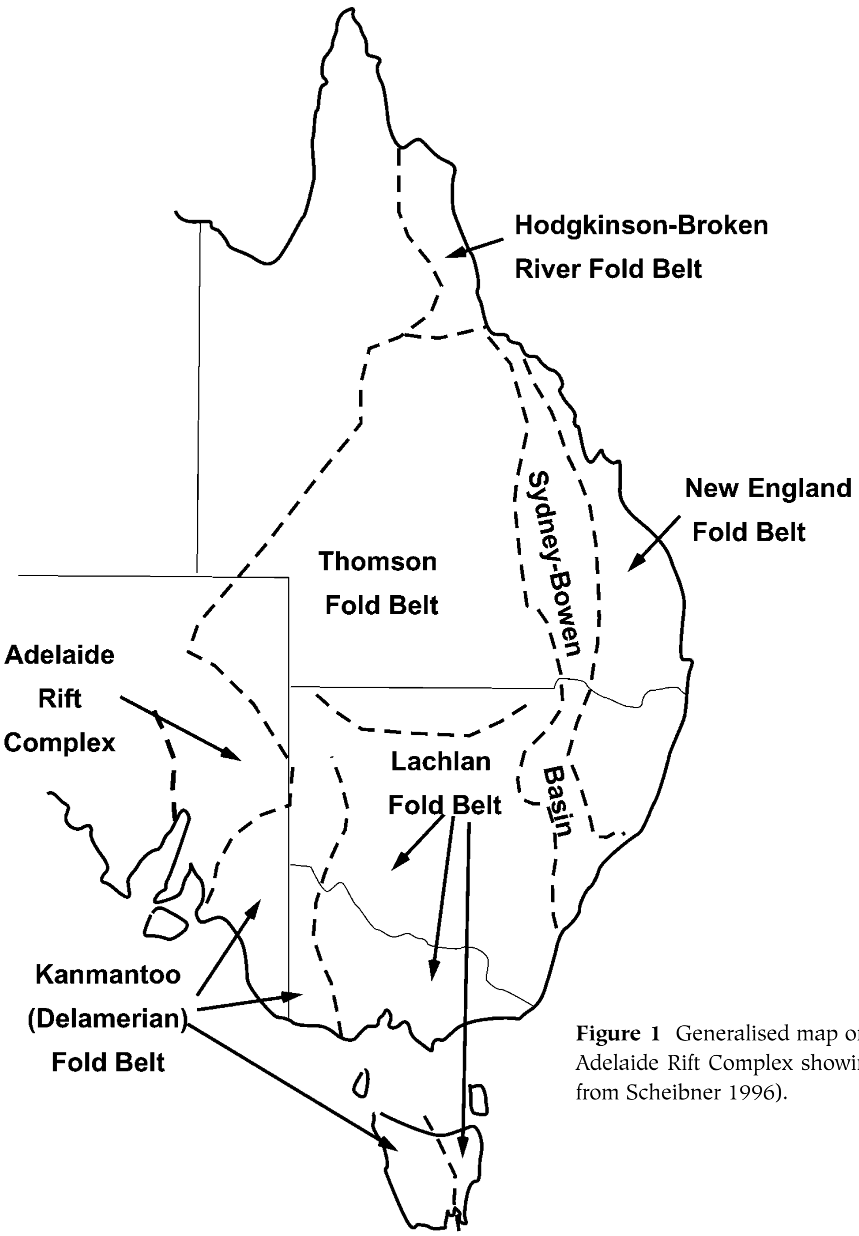


Figure 1 Generalised map of the Tasman Fold belt System and Adelaide Rift Complex showing constituent fold belts (modified from Scheibner 1996).

fold belts has been ascribed by several generations of Australian geologists to the progressive accretion to the Australian Mesoproterozoic nucleus of outboard elements along the southwestern margin of the remarkably long-lived Pacific Ocean (Crook 1980). There is consensus that continent–continent collisions have not been involved in generating eastern Australian continental crust (Coney 1992; Veevers 2000a; Collins in press).

The Tasman Fold Belt System formed by collisions of arcs or microcontinents with continents, a process Scheibner (1996 p. 21) referred to as 'collisional accretion'. Such a mechanism of continental growth is not restricted to the present western Pacific margin. A broadly similar tectonic style is recorded from the Altaids of Asia (Sengor & Natal'in 1996), the Pan-African fold belts of the Arabian Shield (Johnson *et al.* 1987), and the Cordilleran mountain belt of western North America (Coney 1987). Given the global distribution, since at least the start of the Phanerozoic, of fold belts that display all or most of the stratotectonic features of these collisional accretion fold belts, it is clear that they have been important in the growth of continental crust.

Interpreting the eastern Australian fold belts in an actualistic framework relies heavily on an understanding of the temporal evolution of modern western Pacific-type subduction-related settings. Implicit in this 'the present is the key to the past' framework is that the processes and products of the modern western Pacific arc–backarc basin systems (and their relative dispositions?) are akin to those operating at least as far back as 600–500 Ma around the proto-Pacific. Therefore, a broad-ranging knowledge of modern western Pacific geodynamics and the tectonic settings of magma genesis is a prerequisite to successfully unravelling the geological evolution of the Neoproterozoic–Palaeozoic fold belts of eastern Australia. In fact, it could be convincingly argued that the geodynamic evolution of the eastern margin of the Australian plate since *ca* 120 Ma, including the opening of several major and minor ocean basins, and formation of several microcontinents and intra-oceanic arc systems, is simply a continuation of the same regime of tectono-magmatic events that formed the continental crust of Australia over the preceding 500 million years.

In this paper, we first examine the post-120 Ma geodynamic and tectono-magmatic evolution of the eastern margin of the Australian plate between the Tonga–Kermadec–Fiji–Vanuatu–Solomons arc systems and the eastern Australian continental margin. Armed with this information, we then investigate the geological evolution of southeastern Australia since 600 Ma, and identify key components of former plate-margin settings now incorporated into the continental crust of this region. A comparison of the active (and recently inactive) *in situ* crustal elements of the western Pacific east of Australia with the cratonised crust of southeastern Australia lends insight into the processes that generate continental crust, and into crustal growth western Pacific-style.

FRAGMENTATION OF CONTINENTAL CRUST OVER A RETREATING SLAB

Eastern margin of the Australian Plate 120–45 Ma

The nature of the eastern boundary of the Australian Plate between 120 and 45 Ma is poorly understood. However, there is abundant evidence that commencing around 130–120 Ma, the continental crust of the eastern part of Gondwana passed into a long-lived episode of extension and eventual rifting. This led to fragmentation of continental crust broadly parallel with the present eastern coast of Australia, to form three subparallel marginal basins—from west to east, the Tasman Sea, the New Caledonia Basin and the Loyalty Basin (Symonds *et al*. 1996; Auzende *et al*. 2000; Müller *et al*. 2000)—and two elongate microcontinental ribbons, the Lord Howe Rise and the New Caledonia–Norfolk Ridge (Figure 2). The first sign of eventual rifting and ocean opening was the development of a 2500 km-long continental rift along the eastern margin of Australia mainly between 120 and 95 Ma, characterised by immense volumes of largely felsic magmas (Bryan *et al*. 1997). A southern continuation of this continental rift is presently unknown, but its presence along the central and southern Lord Howe Rise is likely given the immense volume of sediments derived from these rocks that was deposited in the Bass and Otway Basins in the Early Cretaceous.

There is insufficient evidence to be sure whether subduction continued east of the developing rift systems between 120 and 45 Ma. Collot *et al*. (1987) have invoked east-dipping subduction west of New Caledonia from 80 Ma to Early Paleocene (*ca* 64 Ma), although there are presently no known rocks of this age with subduction-related signatures to support this model. Sdrolias *et al*. (2002) also argued for a long period of east-directed subduction east of the Lord Howe Rise during the period from around 90 Ma until about 45 Ma. Given the general northward (albeit slow) absolute motion of the Australian plate during this interval, and the major extension and eastward transport of crustal elements originally contiguous with eastern Australia (including growth of at least two marginal basins floored by oceanic crust), it is difficult to conceive the 90–45 Ma plate boundary east of Australia being anything but a convergent boundary.

Tasman Sea–New Caledonia Basin–South Loyalty Basin opening

GEOPHYSICAL RECORD

The New Caledonia Basin lies east of the Lord Howe Rise and west of the Norfolk–New Caledonia Ridge (Figure 2). Most studies suggest that extension in the New Caledonia Basin commenced in the Early Cretaceous and that this basin is underlain by extended continental lithosphere (Etheridge *et al*. 1989; Uruski & Wood 1991; Sdrolias *et al*. 2002), although Sutherland (1999) suggested that the southern part may have a limited amount of oceanic-type crust. Sdrolias *et al*. (2002) have proposed that the New Caledonia Basin formed as a backarc rift behind the Norfolk Ridge volcanic arc above a west-dipping Benioff zone involving subduction of Cretaceous (before 120–100 Ma) Pacific Ocean crust.

Ocean crust accretion in the Tasman Sea commenced around 84 Ma southeast of Tasmania, and propagated northward; by 64 Ma the Coral Sea commenced opening (Gaina *et al*. 1998). All spreading between the Lord Howe Rise (and associated crustal elements) and Australia had ceased by 52 Ma.

Oceanic crust lying between the New Caledonia–Norfolk Ridge and the east-dipping modern subduction zone of the Vanuatu–Solomons arc systems has been referred to as the South Loyalty Basin for that part west of the Loyalty–D'Entrecasteaux Ridge, and the North Loyalty Basin between that ridge and the Vanuatu (New Hebrides) island arc. A large part of the North Loyalty Basin has been subducted beneath the Vanuatu arc, at least over the last 5–10 million years during opening of the North Fiji Basin.

Kroenke (1984) and Bitoun and Recy (1982) suggested that the South Loyalty Basin formed in the Eocene. In contrast, Dubois *et al*. (1973) and Collot *et al*. (1987) argued that the South Loyalty Basin formed as a marginal basin prior to 80 Ma above an east-dipping subduction zone, with a trench west of New Caledonia, implying subduction of oceanic crust of the New Caledonia Basin. New seismic interpretations of South Loyalty Basin sediment fill from the southern part of the basin away from the zone of Eocene emplacement of the New Caledonia ophiolite suggest that the South Loyalty Basin formed by extension and later ocean crust accretion around 120 Ma in the Early Cretaceous (Auzende *et al*. 2000), perhaps contemporaneous with the opening of the New Caledonia Basin. Cretaceous oceanic crust in the South Loyalty Basin is also demanded by the model of Sdrolias *et al*. (2002).

GEOLOGICAL RECORD

Although the nature and age of oceanic crust in the Tasman Sea is well demonstrated by magnetic anomalies, basement has only been sampled at DSDP Site 283, southeast of

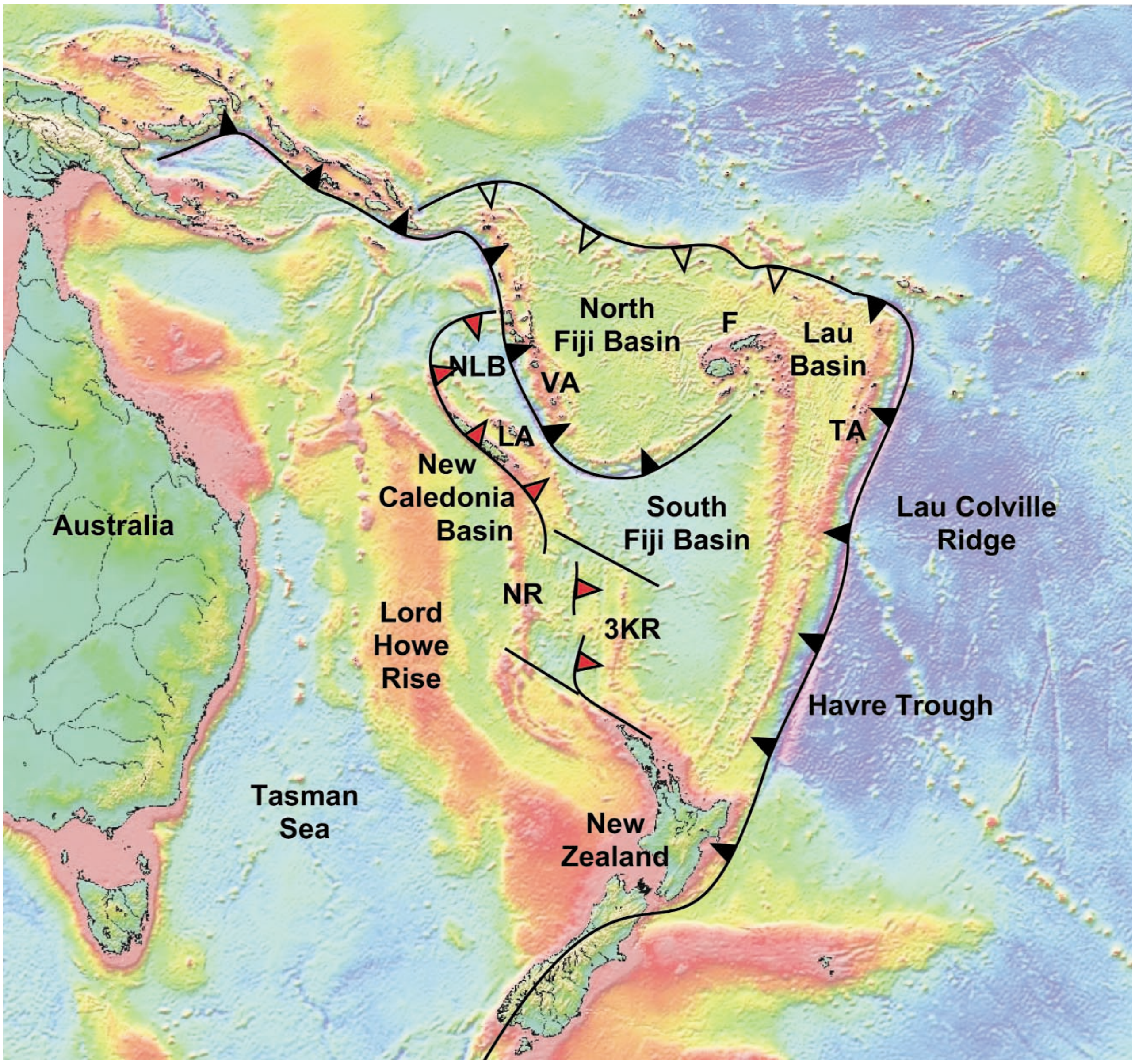


Figure 2 Southwest Pacific east of Australia showing gravity anomalies based on satellite altimetry and major morphotectonic elements. NLB, North Loyalty Basin; LA, Loyalty arc; NR, Norfolk–New Caledonia Ridge; VA, Vanuatu (New Hebrides) arc; TA, Tongan arc; F, Fiji.

Tasmania (Pyle *et al*. 1995). To our knowledge, there is no information on dredged rocks from the floor of the New Caledonia Basin.

Evidence for the nature and age of the earliest oceanic crust along the northwestern margin of the South Loyalty Basin is found in the Formation Des Basaltes of the allochthonous Poya Terrane in New Caledonia. Poya Terrane rocks contain Campanian to Late Paleocene (*ca* 80–55 Ma) microfossils (Aitchison *et al*. 1995; Cluzel *et al*. 2001) and were emplaced as a major allochthon in the latest Eocene (Cluzel 1998; Cluzel *et al*. 2001) in a southwesterly direction over continental crust at the leading edge of the Australian Plate. Basaltic lavas in the Formation Des Basaltes include E-MORB and N-MORB lavas with slight to significant negative Nb anomalies, and εNd values between + 3 and +5 (Eissen *et al*. 1998; Cluzel *et al*. 2001). The latter authors have shown that many Poya Terrane basalts have geochemical affinities with backarc basin basalts. We concur with Cluzel *et al*. (2001) that the Poya Terrane basalts represent early stage magmatism in the volcanic passive margin of a backarc basin. We argue that this basin formed during fragmentation of a Cretaceous arc constructed on thinned continental crust at the leading edge of the Australian Plate.

There is no evidence in the Poya Terrane for extension-related volcanic rocks as old as the 120 Ma sequences identified seismically in the offshore South Loyalty Basin by Auzende *et al*. (2000). There are several possible reasons for this: (i) it is possible that seismic correlations with sequences west of and on New Caledonia are incorrect or inapplicable; (ii) the Early to Middle Cretaceous sedimentary packages identified seismically in the South Loyalty

Basin (Auzende *et al*. 2000) may be early sag and rift phase sequences that significantly pre-date crustal rupture and ocean crust accretion, dated by the Poya Terrane basalts as occurring around 80–70 Ma; and (iii) microfossils indicating a Maastrichtian through Late Eocene age for at least some of the Poya Terrane basalts (Cluzel *et al*. 2001) may be from the sedimentary cappings on post-oceanic crust accretion intraplate seamounts identified among the Poya Terrane basalt piles. For the present time, we regard accretion of oceanic crust in the South Loyalty Basin as being essentially contemporaneous with that in the Tasman Sea.

Further evidence for the existence of Cretaceous oceanic crust in the South Loyalty Basin east of the New Caledonia–Norfolk Ridge is also preserved much further south, in the Northland allochthon in New Zealand. Here, southwest-directed emplacement of Late Cretaceous oceanic basalts onto the North Island occurred around 25–23 Ma (Malpas *et al*. 1992; Mortimer *et al*. 1998; Nicholson *et al*. 2000). The basalts show a compositional range very similar to that of the Poya Terrane basalts, with both N-MORB and E-MORB represented. Some basalts show negative Nb anomalies implying, according to Nicholson *et al*. (2000), generation in a backarc basin close to an island arc. Whether the Northland allochthon represents the terminal phase of a diachronous collision that commenced in the New Caledonia region *ca* 40–34 Ma and swept south over 20 million years, or whether they are two quite unrelated collision events, is presently unknown.

A PLUME-TRIGGERED EVENT

Evidence summarised above indicates that following a substantial period (120–80 Ma) of crustal extension and basin formation within the continental crust of eastern Australia, oceanic crust accretion commenced in the Tasman Sea and South Loyalty Basin at about 85 Ma. Why was basin formation and eventual ocean crust accretion in this easternmost part of Gondwana continental crust so widely distributed, so that instead of a single ocean basin being formed, two subparallel marginal seas and the intervening New Caledonia Basin rift developed?

Evidence for the arrival of a major mantle plume beneath the southern Lord Howe Rise–east Tasmania region very close to 100 Ma is widespread. Lanyon *et al*. (1993), Weaver *et al*. (1994) and Storey *et al*. (1999) showed that a profound change in the isotopic composition of intraplate basalts occurred in Tasmania, New Zealand and the Marie Byrd Land–Transantarctic Mountains region of Antarctica immediately after 100 Ma. All basaltic magmatism since about 100 Ma has a distinct HIMU isotopic fingerprint, indicating massive-scale emplacement of HIMU source mantle beneath this region. No large igneous province has been linked to arrival of this plume head beneath the regional southeastern Gondwana lithosphere, although regional uplift is very well documented (Baker & Seward 1996; O'Sullivan *et al*. 2000; Spell *et al*. 2000).

The role of mantle plumes in initiating crustal extension and ocean opening has been emphasised by White and McKenzie (1989), who demonstrated the occurrence of volcanic passive margins in such plume-triggered rift-ocean settings. Well-known examples include the North and South Atlantic margins. Breakup in both the North and South Atlantic, and on the northwestern margin of Australia, was marked by excess basaltic magmatism at the continent–ocean boundary and the production of seaward-dipping reflector packages along several thousand kilometres of the conjugate passive margins (Planke *et al*. 2000).

Unlike the South Atlantic margin, for example, seaward-dipping reflector packages are presently unknown along the Tasman Sea–Lord Howe Rise margin. The western margin of the Lord Howe Rise is a classic lower plate margin as defined by Lister and Etheridge (1989). It is underlain by a 200 km-wide rift zone characterised by eroded tilt blocks and half-grabens, with west-dipping faults and eastward-rotated fault blocks. The eastern Lord Howe Rise margin is an eroded basement block about 150–200 km wide that resembles an unstructured ribbon-type marginal plateau. Evidence for a volcanic passive margin along the western side of the South Loyalty Basin is present as the Poya Terrane basaltic allochthon. Instead of a single rift developing into an ocean basin South Atlantic-style, extension in the Tasman Sea–Lord Howe region was distributed over a much wider area. We suggest that this exceptionally wide zone of rifting along eastern Gondwana developed in response to rapid rollback of the subducting plate further east. Whether this subducting plate was west-dipping Pacific Plate, or perhaps east-dipping crust being subducted along the eastern side of the New Caledonia–Norfolk Ridge (Sdrolias *et al*. 2002) is presently uncertain. We believe that excess magmatism due to sudden decompression at breakup, which might normally be focused into a developing rift to form seaward-dipping reflector packages along the margins of a single ocean basin, was instead distributed over a large area and led to near-simultaneous formation of the Tasman Sea and South Loyalty Basin. Accretion of oceanic crust ceased abruptly in the Tasman Sea at *ca* 52 Ma, probably in response to the first effects on global plate motions of the India–Asia collision.

Formation of microcontinental blocks: the Müller–Gaina model

Gaina *et al*. (1998) showed that there appear to have been 13 microcontinental blocks involved in the opening of the Tasman Sea. An important aspect of the diachronous peeling off of the long, compound continental crustal slice to form the Lord Howe Rise (and crust further east on the Norfolk Ridge) is that small blocks of continental crust were isolated by unsuccessful rifting events. Notably, the East Tasman Plateau separated from the Lord Howe Rise by 83 Ma, the Gilbert Seamount Complex crustal block by *ca* 77 Ma, and the Dampier Ridge by 64 Ma. All these small crustal blocks were formed by ultimately unsuccessful ridge propagation and the jumping of the spreading ridge to a more favourable location (Gaina *et al*. 2002). Based on the Lord Howe region, and other examples including the Seychelles, Elan Bank (Kerguelen Plateau) and the Wallaby Plateau off northwestern Australia, Müller *et al*. (2001) and Gaina *et al*. (2002) proposed that this ridge-jump model is an important mechanism in the formation of marginal continental ribbons. A key feature in this model is the role of a mantle plume, the existence of which at *ca* 100 Ma is well established for the southern Tasman Sea–Lord

Howe Rise region. As a young continental margin passes close to or over a plume, the inboard flank of the rifted margin is thermally weakened, forcing a ridge jump to this zone of weakness and eventual calving-off of a microcontinental slice. Thus plume–ridge interactions along the eastern Gondwana margin around 100 Ma, coupled with a retreating subducting slab, generated the microcontinental ribbons of the Lord Howe Rise and associated blocks, and the Norfolk–New Caledonia Ridge.

What is the fate of these microcontinental ribbons? To investigate this problem, we need to examine the geodynamic evolution post-100 Ma of this outboard section of the eastern Gondwana margin, directly facing the Pacific Plate, especially with respect to its history of subduction.

SUBDUCTION-RELATED TECTONICS

Subduction before 45 Ma

We have noted above that the enormous extension of eastern Australian lithosphere and formation of marginal seas between *ca* 100 Ma and 45 Ma, occurring adjacent to a boundary between the Australian and Pacific Plates, would appear to demand contemporaneous subduction and slab rollback. A number of workers (Dubois *et al*. 1973; Collot *et al*. 1987) proposed that east-directed subduction of New Caledonia Basin oceanic crust occurred before 80 Ma, with the trench occurring along the western edge of the New Caledonia–Norfolk Ridge. However, this model demands the presence of a Late Cretaceous–Paleocene island arc in the immediate New Caledonia–Norfolk Ridge region for which there is still no evidence (Regnier 1988). Furthermore, there is no convincing evidence that the stretched continental crust of the New Caledonia Basin ever reached the stage of rupturing and accretion of oceanic-type crust, so that we consider subduction of New Caledonia Basin crust to be highly unlikely.

Sdrolias *et al*. (2002) also supported the existence of Cretaceous subduction east of Australia, and argued for west-dipping subduction of Pacific oceanic crust prior to 120–100 Ma, with arc magmatism occurring on the Norfolk Ridge. This was followed by east-dipping subduction between 90 and 45 Ma, with the trench lying west of the Loyalty–Three Kings Ridge. Again, arc-type volcanics of Late Cretaceous to Eocene age are thus expected along the Norfolk and Loyalty–Three Kings Ridges, but to date none have been found.

Based on modern west Pacific arc–backarc basin systems, whether above west-directed (e.g. Tonga, Mariana, Ryukuyu systems) or east-directed (e.g. Vanuatu arc) subduction, extension is always on the upper plate, in part as a consequence of rollback of the subducting plate. This predicts that the Cretaceous–Paleocene extensional setting east of Australia was due to massive rollback of the Pacific Plate, demanding west-directed subduction of Pacific Oceanic crust (Figure 3a). Perhaps slab rollback was so rapid and effective that arc magmatism was dampened or suppressed, and the major magmatic expression of subduction was backarc basin-type ridge-generated oceanic crust of the Tasman Sea and South Loyalty Basin. In the Mariana–West Philippine arc–backarc basin system that has formed over subducting Pacific crust since *ca* 45 Ma, arc magmatism either shut down or diminished remarkably in volume during opening of the Parece Vela–Shikoku backarc basin from 32 to 16 Ma (Crawford *et al*. 1981).

The reason for the apparently poorly represented or absent arc-type magmatic products associated with this Cretaceous subduction is a first-order problem for future studies of the southwest Pacific sea floor.

Subduction initiation at ca 45 Ma?

Numerous previous workers have proposed that subduction initiation around the western Pacific commenced around 45 Ma, as a consequence of plate boundary readjustments following collision of India with Asia (Veevers 2000b). In the Mariana–west Philippine region, subduction initiation generated a nearly 3000 km-long, several hundred kilometres-wide belt of new lithosphere dominated by boninitic lavas and their cumulate counterparts (Crawford *et al*. 1981; Stern & Bloomer 1992). However, recent dating of these rocks indicates eruption ages between 52 and 48 Ma (Cosca *et al*. 1998). The key point is that subduction was perhaps as much as 10 million years before 45 Ma.

There are several models accounting for the production of this massive belt of boninitic lithosphere (Crawford *et al*. 1989; Pearce *et al*. 1992; McPherson & Hall 2001). The key ingredient in any model must be that the partial melting of previously melted, highly depleted lithospheric mantle requires exceptionally hot conditions (>1200°C) at very shallow levels (<50 km) in the subduction zone. Such conditions are not normally present in intra-oceanic subduction zones at depths < 50 km. We prefer a model for boninite generation in the Mariana–Bonin system in which the extra heat required for boninite production derives from subduction of an active spreading centre orientated almost parallel to the trench (Crawford *et al*. 1989), the ridge involved being that on the now-subducted North New Guinea Plate (Seno & Maruyama 1984).

Eocene boninites are also present as thrust slices beneath the main New Caledonia ultramafic ophiolite and in the forearc region of the north Tongan arc (see later). We believe that these formed at the same time as the main Mariana–west Philippine boninitic crustal sections, the extra heat required in this model being due to subduction nucleating along a recently inactive spreading centre in the South Loyalty Basin (Eissen *et al*. 1998). We argue that these boninitic successions represent a key event in the western Pacific region, recording a major change in plate kinematics at *ca* 55 Ma, not 45 Ma, and an eventual shift to more west-directed motion of the Pacific Plate. What is the record of this Eocene subduction event in the southwest Pacific?

Loyalty Ridge–D'Entrecasteaux Ridge

The eastern margin of the South Loyalty Basin is defined by the Loyalty Ridge, a discontinuous bathymetric high formed by well-spaced seamounts that occasionally emerge to form the Loyalty Islands. The Loyalty Ridge trends northwest (Figure 2) before swinging around clockwise to form the east-trending D'Entrecasteaux Ridge that is presently colliding with the mid-segment of the Vanuatu island arc.

Between the Loyalty–D'Entrecasteaux Ridge and the active Vanuatu (New Hebrides) trench is the North Loyalty Basin (Weissel *et al.* 1982).

Islands and seamounts forming the Loyalty–D'Entrecasteaux Ridge show evidence of at least two stages of magmatic activity. Intraplate basalts that were erupted between 11 and 8 Ma cap Mare, one of the southern Loyalty Islands (Monzier *et al.* 1989). In contrast, dredging and ODP drilling has shown that the seamounts further north, including one in collision with the Vanuatu arc, are typical intra-oceanic arc volcanoes, dominated by primitive arc tholeiitic lavas (Maillet *et al.* 1983; Baker *et al.* 1994; Coltorti *et al.* 1994a). ODP drill site 831 on Bougainville Guyot, located at the eastern extremity of the D'Entrecasteaux Ridge, has shown that this was an Eocene arc volcano. Furthermore, rocks drilled beneath middle Eocene ooze at ODP Site 828 on the North D'Entrecasteaux Ridge north of Bougainville Guyot are also primitive arc tholeiitic magmas appropriate for a forearc setting of magma generation (Coltorti *et al.* 1994b). At about 38–37 Ma, magmatism terminated and Bougainville Guyot volcano subsided rapidly (Quinn *et al.* 1994; Baker *et al.* 1994).

On this basis, most authors agree that the Loyalty–D'Entrecasteaux Ridge is probably an Eocene island arc formed over an east- and southeast-dipping subduction zone (Figure 3b) with a trench originally located between the present east coast of New Caledonia and the Loyalty Ridge (Maillet *et al.* 1983; Aitchison *et al.* 1995; Eissen *et al.* 1998; Ali & Aitchison 2000). There are no constraints presently available on the timing of initial arc magmatism on the Loyalty–D'Entrecasteaux arc. If the North Loyalty Basin is regarded as a backarc basin that developed behind this arc, then magnetic anomalies identified as 56–50 Ma by Weissel *et al.* (1982) and Collot *et al.* (1985) suggest that the Loyalty arc was active before 56 Ma. However, in a new interpretation of these anomalies, Sdrolias *et al.* (2002) have proposed instead that they were formed between 43.8 and 35.3 Ma, an age range consistent with the North Loyalty Basin being a backarc basin linked to the Loyalty–D'Entrecasteaux arc. Mid-Eocene volcaniclastic sandstones of arc provenance overlying basaltic basement at DSDP site 286 (Andrews *et al.* 1975) provide a minimum age for this northern part of the North Loyalty Basin crust.

Maillet *et al.* (1983), Aitchison *et al.* (1995), Eissen *et al.* (1998), Ali and Aitchison (2000) and Cluzel *et al.* (2001) have all proposed very similar tectonic models in which the forearc of the Loyalty arc collided in the late-Early to mid-Eocene (around 46–38 Ma) with the Norfolk Ridge, which at that time constituted the thinned leading edge of the Australian Plate. This led to emplacement of the New Caledonia ophiolite. Seismic transects reported by Collot *et al.* (1987) show oceanic crust of the South Loyalty Basin to the west of the Loyalty Ridge rising towards, and being continuous with, the New Caledonian ophiolite. The highly refractory nature of the mantle section of the New Caledonian ophiolite (Prinzhofer 1981; Leblanc 1995), and the occurrence of low-Ca boninites in fault slices beneath the mantle section of the ophiolite (Cameron 1989), strongly support a forearc origin of the main ophiolite on New Caledonia.

Eissen *et al.* (1998) proposed a model for the tectonic development of this region in which east-directed subduction initiated at 50–45 Ma along the active spreading centre in the South Loyalty Basin (Figure 3b). Abnormally high mantle temperatures at the initiation of subduction led to boninite formation in the forearc region of the resultant Loyalty arc (Crawford *et al.* 1989). Continuing east-directed subduction of the western half of the South Loyalty Basin led to growth of the Loyalty–D'Entecasteaux arc and the eventual arrival of the Late Cretaceous volcanic passive margin of this basin at the trench. West-directed emplacement, first of the volcanic passive margin-derived Poya Terrane (Figure 3b), followed by the main New Caledonian forearc-derived ophiolite, occurred around 40–34 Ma (Cluzel *et al.* 2001), terminating subduction and leading to subsidence of arc volcanoes such as Bougainville Guyot. As the age of the youngest magnetic anomaly in the North Loyalty Basin identified by Sdrolias *et al.* (2002) is 35.3 Ma, the cessation of backarc spreading was broadly contemporaneous with shutdown of the arc.

The initial phase of collision involved emplacement of one or more slices of the Late Cretaceous volcanic passive margin at the leading edge of the Australian Plate back southwestward over the adjacent thinned continental crust (Figure 3b). These volcanic-dominated allochthonous sheets are now represented by the Poya Terrane, best exposed along the western coast of New Caledonia. Subsequent thrusting emplaced slices of forearc crust and upper mantle over the Poya allochthon (Eissen *et al.* 1998; Cluzel *et al.* 2001). The uppermost structural slice dominates the ophiolite and is made up largely of uppermost mantle rocks lacking a crustal section (Figure 3b). It may be that the crustal section was sliced off as an earlier allochthon (represented, albeit in very small outcrop areas, by the boninites around Nepoui), to be subsequently overridden by the uppermost mantle section.

Post-collisional tectonic evolution and exhumation of underthrust continental crust

Leading up to, and following emplacement of, the main New Caledonia ophiolite at 40–34 Ma, a foredeep developed in which bioclastic and pelagic sediments were overlain by coarse clastic sediments and olistostromes, contemporaneous with thrust faulting (Aitchison *et al.* 1995; Cluzel *et al.* 2001). Igneous and sedimentary units of the volcanic passive margin and adjacent thinned Early Mesozoic continental crust on the downgoing plate were metamorphosed to eclogite and blueschist facies (Clarke *et al.* 1997) in the subduction zone above the overthickened crust. Exhumation and tectonic denudation of these high-pressure rocks (Figure 3b) occurred in response to post-collisional extension and development of large detachment faults (Rawling & Lister 2000). Sediments produced during this unroofing process and accumulated in basins on either side of New Caledonia are 1–1.5 s (TWT) thick (Auzende *et al.* 2000). Below them, an erosional break between Middle Eocene, and Late Eocene to Late Oligocene strata reflects emplacement of the ophiolite.

There is some evidence that cessation of east-dipping subduction may have swept southward, affecting New Caledonia around 40–34 Ma and as late as 25 Ma when the

3a

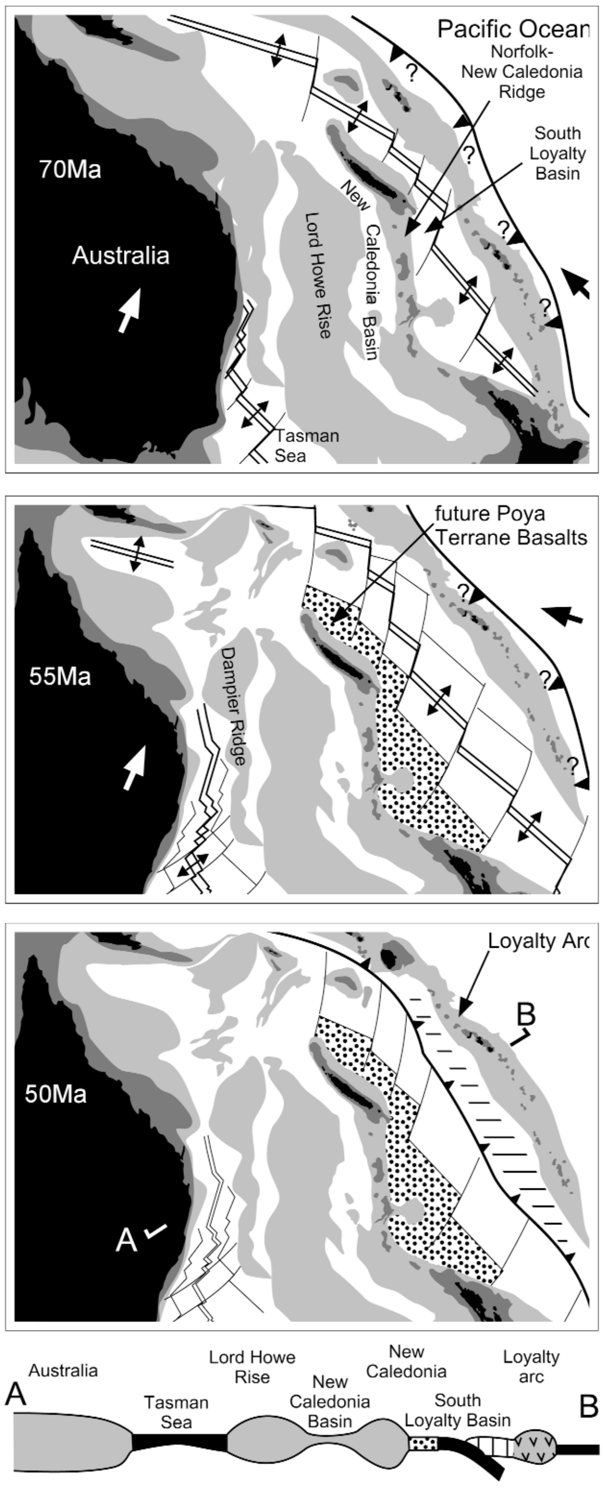
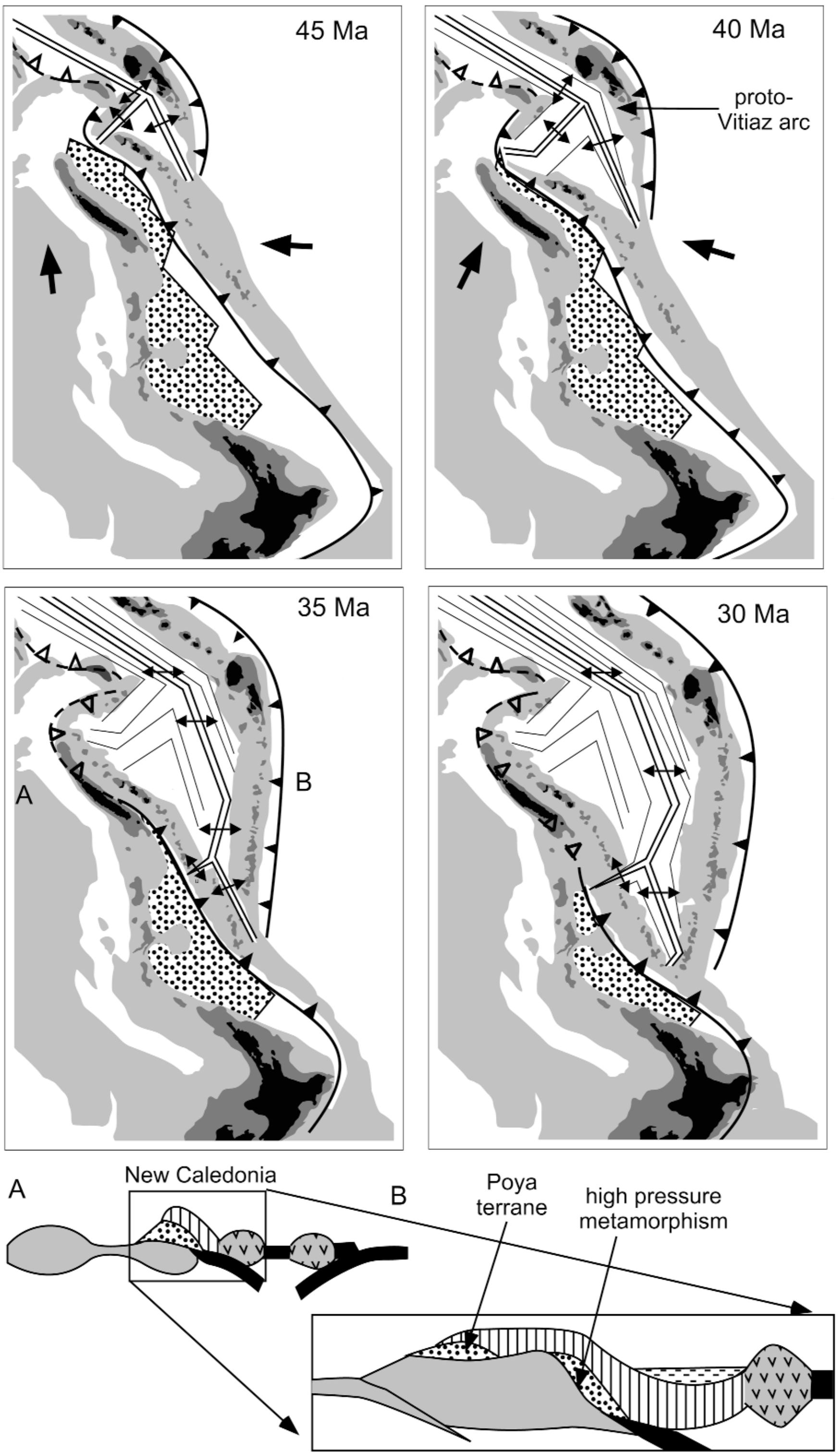


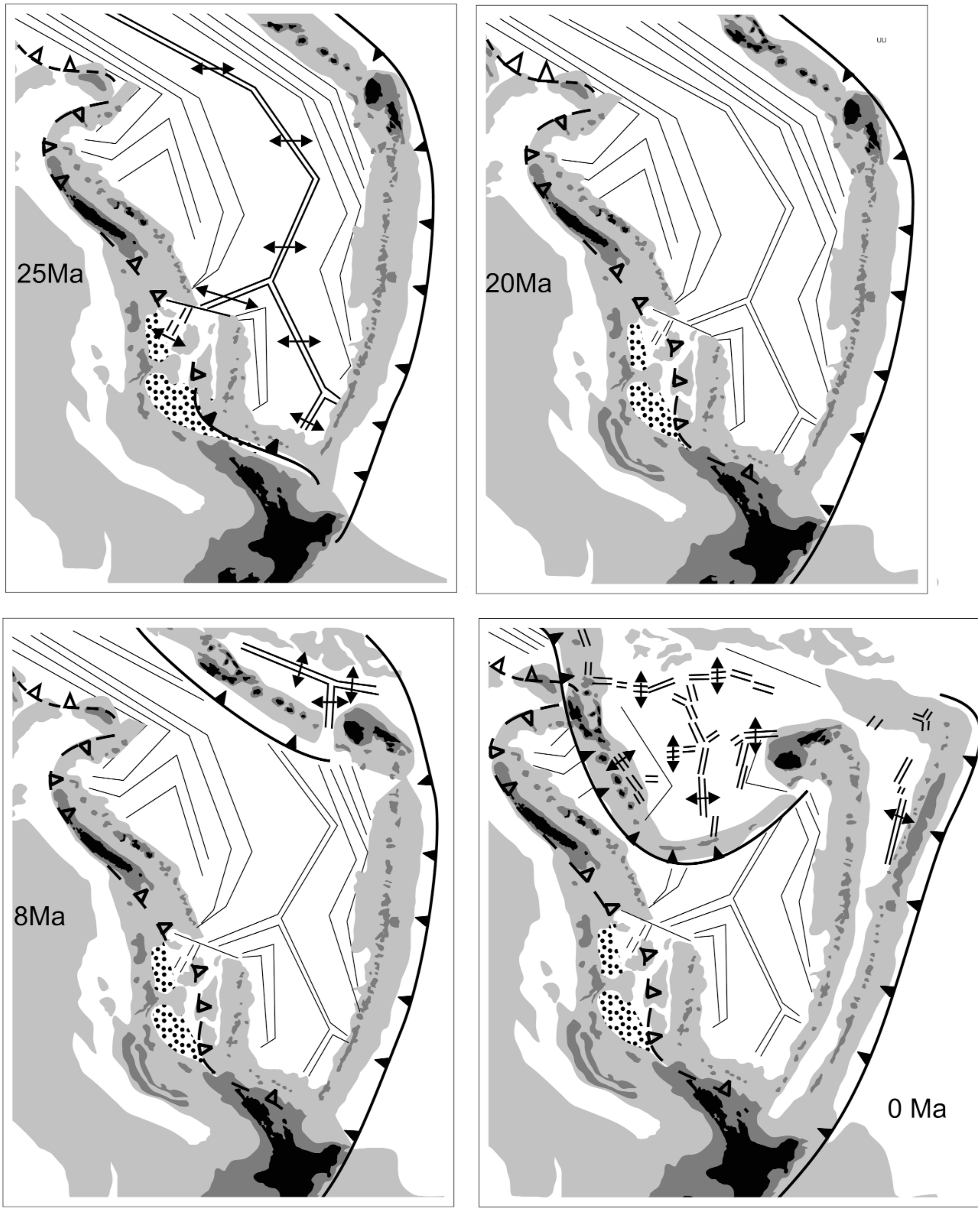
Figure 3 Model for the tectonic evolution of the southwest Pacific from 70 Ma to present. Heavy arrows show approximate plate motion directions (a) Setting at *ca* 70 Ma showing ongoing accretion of oceanic crust in the Tasman Sea and South Loyalty Basin, and crustal extension in the New Caledonia Basin, over a west-dipping subduction zone marked by rapid rollback of the trench. By 50 Ma, northward propagation of the Tasman Sea spreading ridge has detached the Dampier Ridge, and spreading ceased at this time in the Tasman Sea and South Loyalty Basin. East-dipping subduction was initiated between 55 and 50 Ma along the recently inactive South Loyalty Basin spreading centre and generation of boninitic lavas in the forearc of the new Loyalty–D'Entercasteaux arc, which is built upon pre-existing arc crust. (b) At 45 Ma, a major change in global plate kinematics leads to initiation of west-dipping subduction, as opening of the North Loyalty Basin starts to unzip the arc. By *ca* 40 Ma arrival of the western volcanic passive margin of the South Loyalty Basin (eastern margin of the Norfolk–New Caledonia Ridge) at the trench in front of the Loyalty–D'Entrecasteaux arc leads to emplacement of the Poya Terrane nappe, derived from rift tholeiitic basalts of the volcanic passive margin, and subsequent emplace-

3b



ment of boninitic lava slices and sub-boninite refractory uppermost mantle of the forearc region of the Loyalty-D'Entrecasteaux arc to form the New Caledonia ophiolite (see cross-section). The arc subsides by 38 Ma. From 35–30 Ma, the South Fiji Basin continues to grow and the Vitiaz arc extends southward. (c). A western spreading centre limb in the South Fiji Basin propagates into the Norfolk Ridge, detaching the Three Kings Ridge and terminating the east-dipping subduction. Spreading in the South Fiji Basin terminates around 25 Ma, and Cretaceous basalts are thrust southwestward over Northland in New Zealand. Following final collision of the Ontong Java Plateau with the Solomons arc and blocking of this plate boundary around 10 Ma, flipping of the subduction zone to east-dipping occurred, with near-contemporaneous splitting of the Vitiaz arc to form the North Fiji Basin. Volcanism on the Vanuatu arc commences around 8–5 Ma and continued rotation of the Fiji and Vanuatu segments of the Vitiaz arcs occurs via further opening of the North Fiji Basin. Rifting of the Tonga–Kermadec arc starts in the north at *ca* 6–5 Ma with opening of the Lau Basin.

3c



Northland allochthon was emplaced. Whether this diachronous collision commenced further north than the New Caledonia region is unknown at present.

Eocene subduction: regional evidence

We have argued above that in response to a major change in global plate motions at *ca* 55 Ma, east-dipping subduction commenced along the recently inactive spreading centre in the Loyalty Basin generating, in the first instance, primitive boninitic crust and complementary refractory harzburgitic upper mantle. As the subduction system reached steady state, individual arc volcanoes of the Loyalty–D'Entrecasteaux arc were constructed by low-K arc tholeiitic magmas. However, there is strong evidence, best shown by the bend in the Hawaii–Emperor chain, that another major phase of plate reorganisation commenced around 45 Ma, with subduction in the southwest Pacific region changing from east-directed to west-directed polarity. We believe that a new arc system, the Vitiaz arc, formed above this west-dipping subducted Pacific Ocean crust (Figure 3b). Evidence reviewed below supports initiation of the Vitiaz subduction zone around 45 Ma.

Tonga–Kermadec arc

The oldest known rocks in the Tonga–Kermadec arc are 46–40 Ma (Ewart *et al*. 1977; Duncan *et al*. 1985) arc-type lavas occurring below upper Middle Eocene limestones on Eua Island (Tappin & Balance 1994) in the forearc region of the Tongan arc. ODP drilling at Site 841 in the Tongan forearc recovered a thick sequence of low-K arc tholeiitic rhyolites (Bloomer *et al*. 1994), dated by McDougall (1994) at 44 ±2 Ma. Further north in the Tongan forearc, true low-Ca boninitic rocks and associated backarc basin-type basalts of probable Eocene age have been dredged at depths in excess of 4 km (Falloon *et al*. 1998) and have yielded dates between 45 and 35 Ma (Bloomer *et al*. 1998). On available evidence, we cannot be sure whether these boninitic lavas were generated in the same *ca* 55 Ma subduction-initiation event that generated the boninite–refractory-forearc mantle package of the New Caledonia ophiolites further north along the same plate boundary, or whether they represent magmatism associated with the 45 Ma Vitiaz arc subduction-initiation event.

Fiji region

In Fiji, the oldest known rocks are those constituting the Yavuna Group of western Viti Levu (Rodda 1994). Volcanic rocks include common pillow basalts for which very little data are available. Late Eocene ages have been established for overlying limestones (Hathway 1994). Further geochemical and dating studies of the Yavuna Group are required to compare them with other Eocene magmatic suites in the region.

Yavuna Group lavas are intruded by extensive dyke swarms that are dominated by basalts and basaltic andesites, and subordinate felsic dykes, with affinities transitional between arc tholeiites and backarc-basin basalts (Wharton *et al*. 1995), and some lavas transitional to boninite compositions (Gill 1987). These in turn are intruded by the tonalitic-trondhjemitic Yavuna stock, dated at 34.9 ±0.9 Ma by K/Ar on hornblende (Rodda 1994) and 36.6 Ma by U–Pb zircon dating (I. Williams pers. comm. *in* Wharton *et al*. 1995).

Vanuatu arc

There is no record at present of Eocene lavas in the Vanuatu arc, the oldest rocks being arc-type lavas on the Western Belt islands of Santo, Malakula and the Torres Group. These largely basaltic and andesitic lavas are extensively intruded by basaltic to basaltic andesite dyke swarms and by the Navaka Gabbro, which is Ar/Ar dated at 23 Ma (S. Meffre & A. J. Crawford unpubl. data). On the island of Maewo, Carney and Macfarlane (1978) reported Lower Miocene conglomerates that contain reworked Eocene limestones and undated volcanics that show low-K arc tholeiitic affinities which match the Eocene volcanics in Tonga and Viti Levu. Volcanic clasts in these conglomerates are commonly angular and appear to be locally derived (Neef *et al*. 1985). Carney (1985) considered these clasts to be sourced from a reef-fringed Upper Eocene to Lower–Middle Miocene tholeiitic island arc.

New data for the small much-disrupted Basement Complex ophiolite on the island of Pentecost in the Vanuatu arc provide useful information on the composition and age of the basement on which the Vanuatu arc developed. Uplift and emplacement of the ophiolite into overlying Miocene volcanic and volcaniclastic rocks have been linked to collision of the D'Entrecasteaux Ridge with the modern Vanuatu arc at 3–2 Ma (Greene *et al*. 1994). The ophiolitic lavas, dolerites and peridotites have compositional affinities with backarc basin crust and uppermost mantle (Crawford *et al*. 1998). A doleritic dyke from the ophiolite has yielded an Ar/Ar plateau age of 44.9 ±3.0 Ma, suggesting that the early formed part of the Vanuatu arc may have been originally constructed on Eocene basaltic crust of the early North Loyalty Basin. An amphibolitic mylonite and a greenschist facies metadolerite from the Pentecost ophiolite have Ar/Ar plateau dates of 35.5 ±0.6 Ma and 33.4 ±0.8 Ma respectively (S. Meffre & A. J. Crawford unpubl. data). These ages are taken to date an important episode of regional deformation and recrystallisation, presumably that responsible for the cessation of spreading in the North Loyalty Basin at *ca* 35 Ma.

Vitiaz arc

We believe that following emplacement of the New Caledonia ophiolite during arc–microcontinent collision at *ca* 38 Ma, subduction jumped outward to become west-directed and to form the Vitiaz arc, comprising what is now the older parts of the Tonga–Fiji–Vanuatu and Solomons arc systems (Figure 3b). By 35 Ma, this arc had started to split, forming the South Fiji Basin. Although linear magnetic anomalies define a complex pattern of spreading, including two triple junctions (Figure 3c), it is relatively well established that spreading commenced in the South Fiji Basin in the Early Oligocene around 32 Ma and ceased about 25 Ma (Weissel 1981; Davey 1982; Malahoff *et al*. 1982; Sdrolias *et al*. 2002). Cessation of spreading at about 25 Ma coincides with a major reorganisation of plate boundaries in the western Pacific region at this time (Yan & Kroenke 1993; Sdrolias *et al*. 2002). However, continuing subduction along the Vitiaz arc post-25 Ma (Figure 3c) is well supported by numerous radiometric dates on lavas in at least the Vanuatu and Fijian sections of the Vitiaz arc (Greene *et al*. 1994; Whelan *et al*. 1985; authors unpubl. data).

Initial collision of the Ontong Java Plateau large igneous province with the Solomons section of the Vitiaz arc may have occurred as early as 25–20 Ma (Kroenke 1984; Yan & Kroenke 1993; Petterson *et al*. 1997). However, the main episode of blocking and compression relating to this collision was probably between 10 and 5 Ma (Petterson *et al*. 1997; Wessel & Kroenke 2000). This led to a flip in subduction polarity at about 10 Ma, and shortly thereafter (8–7 Ma) to opening of the North Fiji Basin backarc basin (Figure 3c). The distribution and orientation of spreading centres in the North Fiji Basin has been remarkably transient (Auzende *et al*. 1988) with numerous microplates, parallel spreading ridges, several unstable triple junctions, and probably leaky transform faults. During opening of the North Fiji Basin, Fiji has rotated anticlockwise by more than 100° away from a position on the Vitiaz arc originally along strike from the Vanuatu and Tongan arcs. The Vanuatu section of the Vitiaz arc, in contrast, has rotated clockwise (Musgrave & Firth 1999), creating the North Fiji Basin in the manner of opening double gates (Falvey 1975, 1978; Taylor *et al*. 2000). The Hunter Ridge transform, linking Fiji and the southernmost part of the Vanuatu arc along the southern margin of the North Fiji Basin, constituted a short-lived subduction zone between *ca* 7 and 3 Ma as rotation of Fiji forced subduction of crust in the northeast South Fiji Basin (Figure 3c). Rocks dredged from the embryonic Hunter Ridge arc include boninitic and arc tholeiitic lavas (Crawford & Verbeeten 2000). East-directed subduction beneath the Vanuatu arc has probably been active since the subduction zone flipped at about 10–8 Ma, although no reliable radiometric dates on lavas in the active arc are >5 Ma (Macfarlane *et al*. 1988).

The Lau Basin–Havre Trough is a well-defined backarc system (Figure 2) that has split the length of the Miocene Lau–Colville Ridge section of the former Vitiaz arc. Subduction and consequent backarc-basin opening is fastest in the north, where a complex triple junction exists; in the south, major extension of arc crust has occurred, but steady-state accretion of oceanic crust may not have commenced. Rifting of the Lau–Colville Ridge is considered to have started around 6 Ma (Clift & Dixon 1994).

New Zealand region

Collision and ophiolite emplacement occurred in the New Caledonia region between 40 and 34 Ma. More than 2000 km further south, similar Cretaceous oceanic crust was thrust westward on to Northland in New Zealand, forming the extensive Northland Allochthon (Malpas *et al*. 1992; Nicholson *et al*. 2000) with the date of emplacement being 25–23 Ma.

Mortimer and Herzer (2000) reported sillimanite–garnet schists from Cavalli Seamount, immediately northeast of the tip of Northland and noted that Ar/Ar data suggest these derive from a Paleocene protolith rapidly exhumed at 25–20 Ma. Mortimer *et al*. (1998) also reported a 26 Ma 'biotite-rich' volcanic breccia dredged from the western slope of the Three Kings Ridge, and far more widespread Early Miocene (22–18 Ma) potassic lavas, including shoshonites, in a broad region of the sea floor north of Northland and along the eastern side of the Three Kings Rise. The calc-alkaline to shoshonitic nature of these Early Miocene volcanics in Northland, on the submarine Northland Plateau, and on the Three Kings Rise, has been linked to contemporaneous west-dipping subduction by Mortimer *et al*. (1998) and Mortimer and Herzer (2000). In contrast, we suggest that these lavas represent either magmatism associated with the final phase of east-dipping subduction which had been progressively shutting down from north to south (Figure 3c), or post-collisional magmatism forming seamounts in a regional extensional setting following emplacement of the Northland allochthon and rapid rollback of the subducting slab.

Summary

As a result of regional extension along the eastern margin of Gondwana since *ca* 120 Ma, several microcontinental ribbons were calved off Australia and were separated by intervening marginal ocean basins. This is presumed to have occurred over a slab of Pacific oceanic crust which was rapidly rolling back eastward. A major reorganisation of plate boundaries at *ca* 55 Ma, probably induced by India–Asia collision, led to the initiation of subduction along former spreading centres in the easternmost of these marginal basins and the generation of boninite-dominated lithosphere along the new convergent plate boundary. Further subduction led to construction of the Loyalty–D'Entrecasteaux arc behind the boninitic forearc terrane. Arrival at the trench of the volcanic passive margin of the Norfolk Ridge microcontinental ribbon led to limited underthrusting of the passive margin crust, and consequent westward emplacement of several allochthonous sheets around 38 Ma. The first and lowest of these sheets represents basaltic lavas and associated rocks of the underthrust volcanic passive margin. A second sheet, very poorly exposed, consists of one or more slices of boninitic lavas representing the subduction-initiation forearc package. The highest and thickest slice is dominated by the uppermost mantle section that underlay the boninitic forearc lavas sequences. Following crustal thickening associated with emplacement of these elements, an extensional regime developed that culminated in exhumation and denudation of the underthrust continental crust, some of which shows eclogite and blueschist facies assemblages. Locking of this plate boundary, and a change of spreading direction of the Pacific Plate around 45 Ma, led to outward stepping of the subduction zone to the position of the Vitiaz arc and a new cycle of arc growth, arc splitting and backarc-basin formation.

This *ca* 38 Ma western Pacific-style arc–microcontinent collision produced a narrow orogen, probably less than 100 km across. Post-collision magmatism is not known in the New Caledonia–Loyalty Basin region and probably did not form because the subduction zone stepped Pacific-ward several hundred kilometres. In contrast, more than 1000 km further south a collisional event at 25 Ma emplaced similar Cretaceous oceanic crust onto Northland. Although poorly understood at present, extension tectonism in the offshore Northland region following the collision led to the generation of high-K calc-alkaline and shoshonitic magmas between *ca* 23 and 18 Ma, and these formed seamounts in the southern section of the South Fiji Basin and on Northland.

LATE NEOPROTEROZOIC AND EARLY PALAEOZOIC SOUTHEASTERN AUSTRALIA

600–570 Ma mafic volcanic rocks: a volcanic passive margin

The geological history of southeastern Australia between the Late Neoproterozoic (*ca* 600 Ma) and the Early Ordovician (*ca* 490 Ma) records a cycle of continental rifting and ocean opening (*ca* 600Ma), subduction (starting *ca* 515 Ma), and arccontinent collision (Crawford & Berry 1992), with important post-collisional extension, magmatism (500 Ma), exhumation of underthrust Neoproterozoic continental

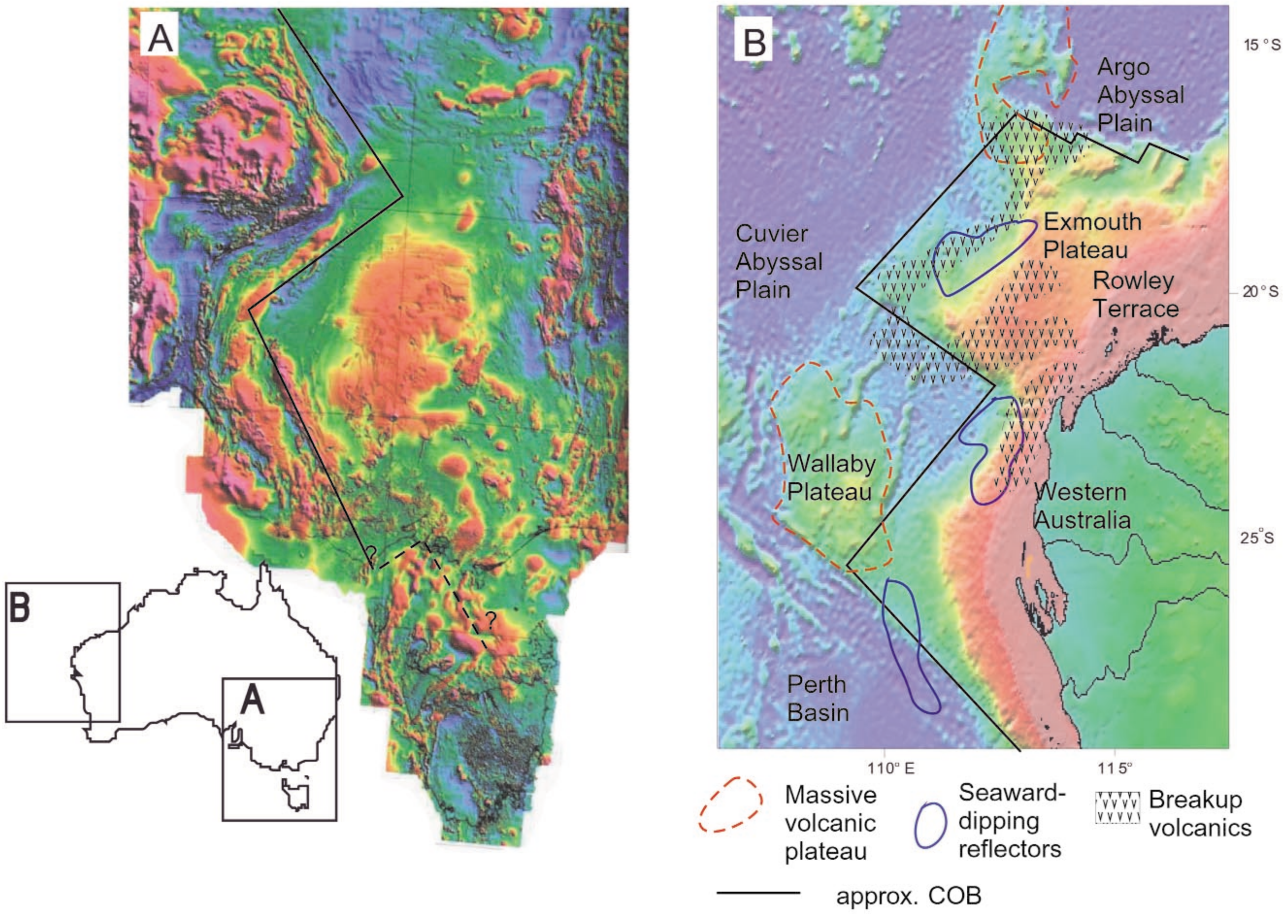


Figure 4 (a) Regional magnetic image of southeastern Australia, showing prominent linear anomalies in western New South Wales, western Victoria, eastern South Australia and in Bass Strait (from VandenBerg *et al.* 2000). (b) Map of the northwestern margin of Western Australia at the same scale as (a) showing orthogonal nature of the breakup margin and distribution of seaward-dipping reflector packages and breakup volcanics (modified from Direen & Crawford 2002).

crust, and molasse deposition. The dimensions, orientation and duration of these events bear a striking similarity to events east of Australia from 100 Ma until the present, and argue for very similar processes having affected the eastern margin of Australia from 600 Ma until the present.

Continental breakup and opening of the palaeo-Pacific has been well defined as occurring between 600 and 560 Ma (Crawford *et al.* 1997; Veevers *et al.* 1997; Foden *et al.* 2001). Abundant evidence occurs in westernmost New South Wales and Victoria, as well as on King Island and in western Tasmania, for the existence of a 600 Ma volcanic passive margin along this section of eastern Gondwana. Rift tholeiitic pillow basalts up to 4 km thick occur along the Koonenberry Belt in western New South Wales (Crawford *et al.* 1997; Crawford & Direen 1998; Direen 1999), and picritic lavas sequences in which liquid compositions had >15% MgO are known from King Island (Waldron & Brown 1993; A. J. Crawford unpubl. data) and western Victoria (Crawford 1997; Direen 1999), and are associated with magnetic anomalies up to several hundred kilometres long (Figure 4a). Carbonate sequences dated around 600 Ma (Calver & Walter 2000) occur below and above the rift tholeiite package in the Smithton Trough in northwestern Tasmania, and at this location record an episode of crustal extension and rift magmatism that probably terminated before rupturing of the continental crust (Crawford & Berry 1992). The axis of extension and magmatism is presumed to have jumped eastward and been successful in opening an ocean basin at 600–570 Ma and in doing so, rifting off at least one continental block or microcontinental ribbon (Crawford & Berry 1992; Crawford & Direen 1998). The presence of high-temperature picritic lavas suggests volcanic passive margin development being triggered by a mantle plume (White & McKenzie 1989). We note the proposition put by Williams and Gostin (2000) that the remarkable submarine canyons cut into Neoproterozoic strata in the Adelaide Rift Complex may be due to regional uplift accompanying emplacement of a mantle plume beneath this region post-590 Ma but before the Precambrian–Cambrian boundary (*ca* 544 Ma).

In this developing Late Neoproterozoic rift, the orientation of spreading centre segments is believed to have been northwest–southeast (Figure 4a), in keeping with the current distribution of major rift volcanic piles on the east-facing volcanic passive margin formed in this rifting event. The northeast–southwest-trending southern margin of the Willyama Block is therefore taken to be a former transform margin at breakup. A second breakup transform may have

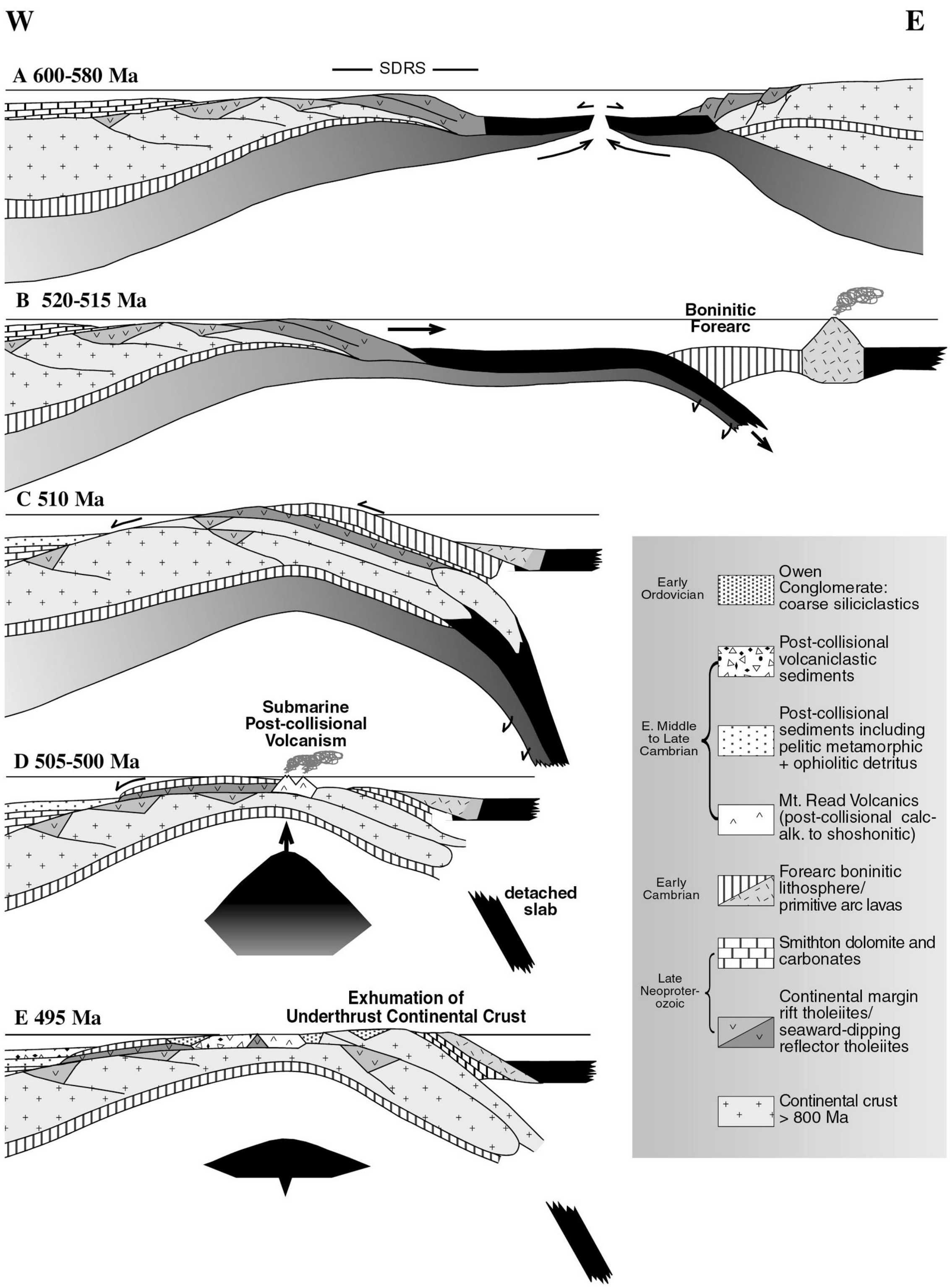


Figure 5 Figure 5: Hypothetical tectonic development of the Tasmanian section of southeastern Australia shown as crustal cross-sections between 600 and 480 Ma. (a) *ca* 600 Ma: plume-triggered rifting at *ca* 600 Ma produces an east-facing volcanic margin with thick seaward-dipping reflector packages (SDRS). (b) *ca* 520–515 Ma: east-dipping subduction commences, and produces a boninitic forearc lithospheric section and a primitive intra-oceanic arc. (c) *ca* 510 Ma: arc–continent collision leads to emplacement of allochthons first of SDRS-type volcanic passive-margin basalts, which are overridden by nappes of forearc-derived boninitic lithospheric sections, leading to collapse of the margin and development of the Dundas Trough foreland basin. (d) 505 Ma: the new crustal collage had commenced to collapse due to post-collisional extension, and Mt Read Volcanics post-collisional lavas were erupted in the half-graben formed between *ca* 505 and 497 Ma. (e) *ca* 495 Ma: continued extension led to exhumation of the underthrust continental crust of the volcanic passive margin, and the production of large amounts of coarse proximal siliciclastic molasse (Owen Conglomerate) that filled grabens formed along the collision zone.

been located between western Victoria and Tasmania to account for the eastward shift in the magnetic trends, corresponding to major Neoproterozoic basaltic piles, going from western Victoria to King Island and Tasmania. The scale and orthogonal nature of the Neoproterozoic volcanic passive margin match well the volcanic passive margin of Western Australia that developed between *ca* 155 and 125 Ma (Figure 4b).

Subduction, ocean closing and post-collisional tectono-magmatic events

Dating the initiation of subduction in this new ocean basin is difficult. Boninitic lava –cumulate piles in western Tasmania and Victoria have yielded Middle Cambrian ages (515–510 Ma: Turner *et al.* 1998). These are considered to have been generated by subduction of a spreading centre, or subduction initiation at a spreading centre, and to have formed the forearc region of an island arc (Figure 5a, b) that was constructed by ongoing subduction (Crawford & Berry 1992). Collision at 510–505 Ma between the 600 Ma east-facing volcanic passive margin and the forearc region of this intra-oceanic arc led to west-directed emplacement of one or more extensive allochthons of forearc-derived boninites and associated low-Ti tholeiitic basalts on to the volcanic passive margin (Figure 5c). At almost every location where the contact is exposed in western Tasmania the ophiolitic rocks are thrust over rift tholeiitic basalts of the Crimson Creek Formation and the 600–570 Ma volcanic passive margin sequences. Faulted contacts with older units, and significant changes in the geochemistry of these basalts at different locations (Brown & Jenner 1989; A. V. Brown & J. L. Everard pers. comm. 2000), suggest that the rift tholeiitic package may also represent one or more allochthonous sheets that were emplaced just before, and were overridden by, the ophiolitic allochthon (Figure 5d). The relative orientation and geological results of this collision are remarkably similar to those that occurred almost 500 million years later during the collision of the Loyalty forearc with the volcanic passive margin of the eastern side of the Norfolk Ridge. In both cases, the orogen formed by the collision was probably less than 100 km wide.

Extension of the newly assembled Middle Cambrian crustal collage led to formation of half-grabens into which, in Tasmania, the Mt Read Volcanics were erupted around 500 Ma almost entirely in a submarine setting (Figure 5d). Further extension in the Late Cambrian led to gradual exhumation of passive-margin crust, some of it bearing eclogite–whiteschist assemblages, from beneath overthrust boninitic allochthons (Crawford & Berry 1992; Meffre *et al.* 2000), so that outcrop of the Mt Read Volcanics was effectively limited to the western and northern margin of exhumed crystalline crust (Figure 5e). Rapid erosion of the actively emergent crystalline crust led to deposition of coarse proximal siliciclastic molasse (Owen Conglomerate and correlates), in part unconformably above the Mt Read Volcanics.

Direct correlates of the western Tasmanian boninitic allochthons and post-collisional Mt Read Volcanics are known across Bass Strait in Victoria. Boninitic lavas of Cambrian age occur in all three major greenstone belts in Victoria and are well exposed in the Heathcote (Crawford & Cameron 1985) and Mt Wellington greenstone belts (Figure 6), particularly at Howqua (Crawford 1982). Following west-directed emplacement in a Middle Cambrian arc–continent collision, these greenstones now form the lowest part of major west-dipping listric thrust duplexes emplaced during east-vergent Late Ordovician or Early Silurian deformation (Fergusson *et al.* 1986; Gray & Foster 1997). Boninitic lavas occur at the base of the lava piles, are overlain by black shales, and followed by up to 1.5 km thickness of backarc basin-type tholeiitic pillow basalts. Conformably overlying the basalts are occasional cherts and a thick pile of Ordovician quartz-rich turbidites that extend across most of the Lachlan Fold Belt from western Victoria to east of Canberra. The Moyston Fault just east of the Grampians in western Victoria is believed to delineate the western edge of the Lachlan Fold Belt and its contact with the Delamerian Fold Belt (VandenBerg *et al.* 2000).

Post-collisional calc-alkaline volcanics with U–Pb zircon dates close to 500 Ma and akin to the Mt Read Volcanics occur in the Stavely greenstone belt in western Victoria (Figure 6) and are best known from the Mt Stavely Volcanic Complex (Crawford *et al.* 1996; Crawford *et al.* in press), which outcrops south of the Grampians and consists of fault-bounded but internally undeformed blocks of volcanic and volcaniclastic rocks. Detrital zircons in a volcaniclastic rock and magmatic zircons in a dacite of the Mt Stavely Volcanic Complex have yielded crystallisation ages of 501 ±9 Ma and 495 ±5 Ma respectively (Stuart-Smith & Black 1999). Cambrian calc-alkaline volcanics with medium- to high-K affinities also occur as thrust slices along the eastern margin of the Melbourne Zone at Jamieson and Licola (Figure 6). In terms of age and geochemical composition, these lavas are strikingly similar to the Mt Read Volcanics of western Tasmania (Crawford *et al.* in press) and they have been interpreted to represent erosion windows through to an older basement of Tasmanian affinity, named the Selwyn Block (VandenBerg *et al.* 2000).

The Selwyn Block is interpreted to extend from Tasmania northward under Bass Strait and to occur under most of the Melbourne Zone in Victoria (VandenBerg *et al.* 2000). VandenBerg *et al.* (2000) showed the Selwyn Block as a microcontinental block probably some 150–200 km wide, and lying 400–500 km east of the eastern edge of the Delamerian Fold Belt in the Late Cambrian, separated from it by Cambrian (backarc basin) oceanic crust represented by the Cambrian tholeiites in the Heathcote and Mt Wellington greenstone belts. By implication, the Selwyn Block should be the microcontinental block rifted from cratonic Australia during the *ca* 600 Ma rifting event.

Following the Delamerian deformation in the Late Cambrian, the subduction zone is believed to have jumped outboard, east of the Selwyn Block microcontinent, to form the Early Ordovician to Late Ordovician Macquarie Arc (Glen *et al.* 1998). Arc-type lavas as old as 480 Ma are recorded from the Ordovician volcanic Parkes–Narromine belt (Butera *et al.* 2001), which trends southward into northeastern Victoria around the headwaters of the Murray River. A detailed discussion of the subsequent Siluro-Devonian tectonic development of the Lachlan Fold Belt is beyond the scope of this paper. However, key features

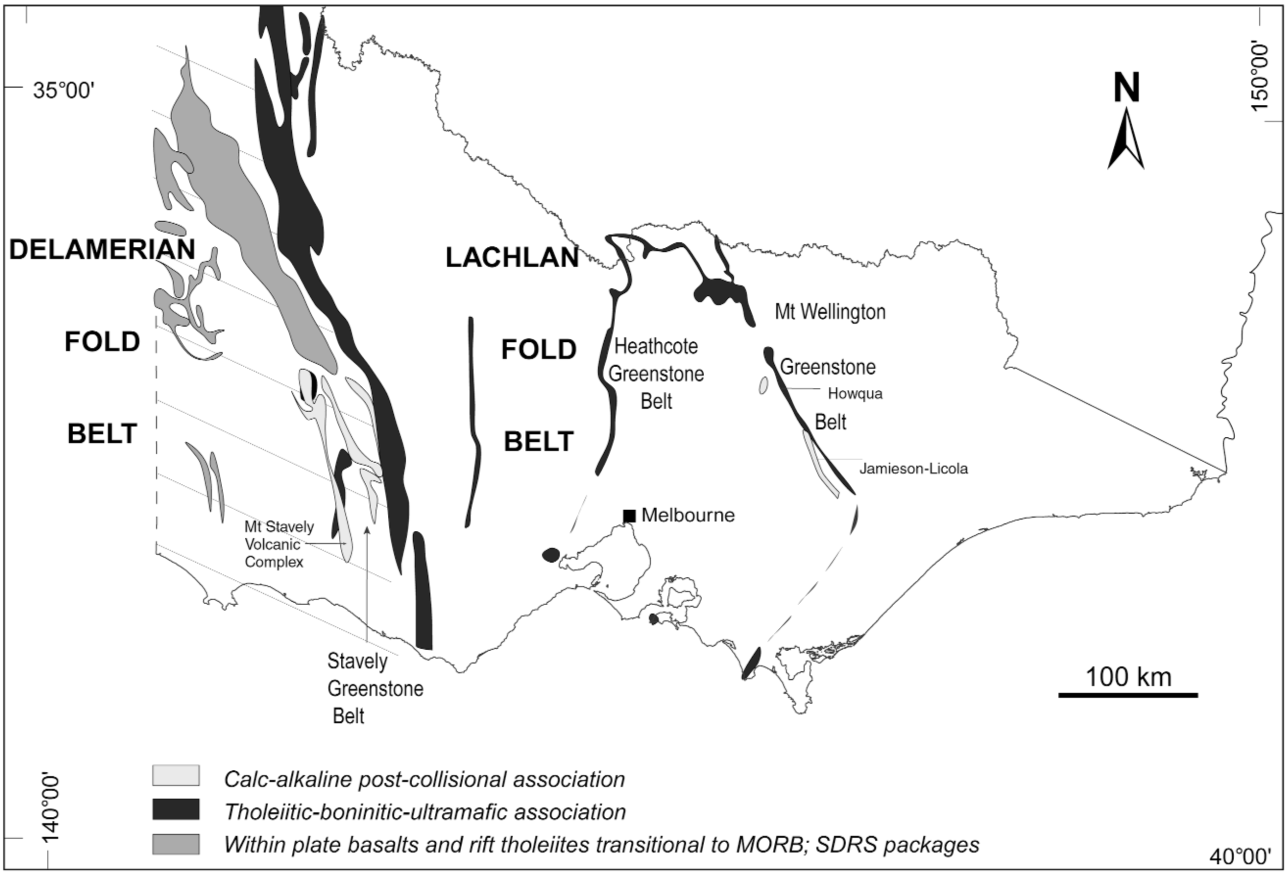


Figure 6 Map of Victoria showing distribution of the Cambrian and Late Neoproterozoic greenstones across the State, divided into Late Neoproterozoic rift-related basalts, allochthonous tholeiite–boninite packages of Early to early-Middle Cambrian age, and late-Middle Cambrian (*ca* 500 Ma) post-collisional calc-alkaline (mainly andesitic) lavas (modified from VandenBerg *et al*. 2000).

worth noting include the development in the Early Silurian through Middle Devonian of subparallel elongate troughs within the Macquarie Arc, including the Cowra and Hill End Troughs, and the closing and deformation of these troughs during the Middle and Late Devonian (Tabberabberan Orogeny). Crust beneath the Snowy Mountains south of Canberra is up to 52 km thick (Finlayson *et al*. 1980), and suggests that a collision–accretion event, involving the extended Macquarie Arc and another microcontinental block, may have been responsible for the Tabberabberan deformation.

Collins (in press) has drawn attention to the intermittently extensional nature of the Lachlan Fold Belt, which he referred to as an extensional accretionary orogen. He argued that such orogens initiate at intra-oceanic convergent margins and grow during slab rollback, by magmatic and protracted turbiditic sediment additions, in the backarc region and in the arc and accretionary prism. Evidence of large scale ‘terrane accretion’ is lacking and Collins (in press) attributes transient localised compressional events to the arrival and temporary blocking effects of ‘oceanic plateaus’. We agree with many aspects of Collins’ model, but argue that for the Tasman Fold Belt System, the ‘oceanic plateaus’ are in fact microcontinental ribbons calved from cratonic Australia probably during 600–570 Ma rifting, in similar manner to the way in which the Lord Howe Rise and Norfolk–New Caledonia Ridge were detached from eastern Australia between 120 and 80 Ma.

New England Fold Belt

A similar sequence of convergent margin development and episodic margin fragmentation and accretion of arcs to eastern Australia occurred between the Devonian and Late Mesozoic with resulting development of the New England Fold Belt (Murray 1987; Aitchison *et al*. 1994; Holcombe *et al*. 1997; Murray *et al*. in press). Included are continental margin arc terranes that presumably formed upon the outboard margin of the recently cratonised Lachlan Fold Belt, and tectono-stratigraphic domains more typical of intra-oceanic arc systems, such as the Gympie ‘terrane’. Calc-alkaline magmatism of Late Devonian–Early Carboniferous age extends up to 400 km inboard of the convergent plate margin, implying a low angle of subduction beneath the continental margin arc at that time. In a recent study of the Gympie Terrane, Sivell and McCulloch (2001) suggested that it formed as a primitive island arc, ‘possibly within pre-existing New England Orogen accretionary complex elements’, during Early Permian extension accompanying eastward trench rollback and steepening of the Benioff zone. The Sydney–Gunnedah–Bowen Basin system was also initiated in the Early Permian by a major extensional event that affected the New England Fold Belt. From the mid-Permian onwards, most of the New England Fold Belt was a major foreland thrust belt that developed in a backarc setting behind the convergent plate boundary and west-dipping subduction zone that had migrated to the east (Korsch & Totterdell 1995). Much of the major Mid to Late Triassic deformation was ascribed by Harrington and Korsch (1985) to the arrival and docking of the Gympie Terrane with the New England Fold Belt, although Sivell and McCulloch (2001) have suggested that accretion of this arc to the continental margin occurred in the Early to Middle Triassic.

Despite ongoing uncertainty relating to the tectonic setting of a number of key elements of the New England Fold Belt (Bryan *et al*. 2001; Murray *et al*. in press), there is little doubt that this orogen developed over a period of more than 150 million years by fragmentation of the continental margin during repeated extensional events, growth of new island arcs with accretionary prisms outboard of this margin, and their eventual return and accretion on to the continental margin. In similar fashion, fragments of this orogen have been split off Australia during Tasman Sea opening and are now located on the Norfolk Ridge (Black 1996; Meffre *et al*. 1996), the Lord Howe Rise, and in New Zealand (Bradshaw 1989; Cawood 1984).

CONCLUSIONS AND IMPLICATIONS FOR RODINIA BREAKUP

A review of the tectonic development of the southwest Pacific east of Australia from 120 Ma until the present has shown that the products of tectono-magmatic processes occurring in this region are remarkably similar to rocks making up the 600–350 Ma southern section of the Tasman Fold Belt System. This implies that very similar processes operated from 600 Ma until now to produce these rocks and assemble the fold belts. Key processes include: (i) the calving-off of elongate microcontinental ribbons probably induced by one or more mantle plumes; (ii) the initiation of subduction and generation of boninitic magmas, probably at a former spreading centre, as a result of a major change in global plate kinematics and reorganisation of plate boundaries; (iii) rapid passage from boninitic magmatism in an extensional setting, to backarc-basin spreading, and eventual growth of an intra-oceanic arc over a steeply dipping and rolling-back slab; (iv) collision of the boninitic forearc of this new arc with a volcanic passive margin of a marginal ocean basin produced during calving-off of the microcontinental fragment(s); (v) emplacement of allochthonous slices, first of volcanic passive-margin basalts, then of forearc-derived boninitic rocks and complementary uppermost mantle, back over the passive margin crust; (vi) locking of the plate boundary and extension-related exhumation of the underthrust volcanic passive-margin elements, with high-pressure metamorphic assemblages; (vii) jumping of subduction, probably with a polarity reversal, outboard towards the Pacific and the creation of a new island arc; and (viii) in response to ongoing trench rollback, cycles of arc splitting, backarc-basin opening, and the construction of new arcs on the remnants of the former arc or on new backarc-basin crust, as exemplified by the North Fiji Basin and Lau Basin.

The Lachlan Fold Belt and New England Fold Belt in southeast Australia can be interpreted in the above framework, implying that the processes active in the southwest Pacific from 120 Ma have apparently been happening along this margin of the Pacific Ocean since at least 600 Ma. Apart from the immense Ontong Java Plateau colliding with the Solomons since *ca* 10 Ma, there is no convincing evidence that any component of either fold belt originated very far from the convergent margin east of Australia. Most or all of the terranes recognised by previous workers in the Lachlan and New England Fold Belts are probably 'native' terranes to the region, having developed within or adjacent to the eastern margin of Australia.

It has been argued that eastern Australia–Antarctica was juxtaposed against Laurentia prior to *ca* 750 Ma, together constituting the supercontinent Rodinia (Dalziel 1991; Hoffman 1991; Moores 1991; Powell *et al*. 1994). Precisely which part of Laurentia lay against eastern Australia is uncertain with candidates including western Canada (Moores 1991), western USA (Burrett & Berry 2000), Mexico (Wingate *et al* 2001) and South China (Li *et al*. 1995). Powell *et al*. (1993, 1994) and Powell (1998) suggested that breakup between Laurentia and Australia–Antarctica happened shortly after 750 Ma and that it occurred along the Tasman Line, and Preiss (2000) considered that breakup occurred at *ca* 700 Ma, closely associated with the Sturtian glaciation. Wingate and Giddings (2000) presented evidence indicating that breakup occurred prior to 755 Ma. However, we are confident that all available geological evidence indicates that the Tasman Line reflects a disrupted and reworked 600–570 Ma breakup margin (Crawford & Direen 1998). If there was a breakup event some 100-150 million years earlier along this same margin, the magmatic evidence of this breakup is dismayingly small, and evidence of the breakup has been reworked or overprinted by the later breakup event.

Strongly influencing thinking regarding the postulated 750–700 Ma age for breakup is the evidence for regional downwarping, intermittent basaltic magmatism and discrete episodes of rifting to form the Adelaide Rift Complex (Preiss 2000). Early extension is dated at 827 ±7 Ma by the northwest-trending Gairdner dyke swarm and the Wooltana and Beda Volcanics. Later rifting (*ca* 700 Ma) is hypothesised to have shifted eastwards and has been interpreted as sag-phase deposition associated with continental separation (Preiss 2000). There is no magmatic expression of this hypothetical breakup.

We suggest that while palaeomagnetic evidence is inconclusive, and geological tests of Australia–Laurentia connections remain arguable, it is worth considering another possibility. That is, that eastern Australia (including its outboard elements such as the Lord Howe Rise) has faced the palaeo-Pacific Ocean since at least 600 Ma and was never a conjugate continental margin (i.e. joined) to some element of Laurentia or South China.

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